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Stratigraphic and geochemical expression of Barremian–Aptian global climate change in Arctic Svalbard

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ABSTRACT

Significant changes in global climate and carbon cycling occurred during the Early Cretaceous. This study examines the expression of such climatic events in high-latitude Svalbard together with the stratigraphic utility of carbon-isotope stratigraphies. Isotopic analysis of fossil wood fragments (from the Rurikfjellet, Helvetiafjellet, and Carolinefjellet formations, Festningen, Spitsbergen) record a distinctive pattern including a negative isotope excursion preceding a positive event, correlatable with the global early Aptian isotope event. Our carbon-isotope profile improves the stratigraphic correlation and relative dating of the succession. We show that the upper part of the Helvetiafjellet Formation was deposited during the early Aptian, and not the late Barremian, as previously thought. Furthermore, we estimate an age for the abrupt contact of the Rurikfjellet Formation with the overlying Helvetiafjellet Formation (associated with a pulse of igneous activity) to be ca. 129 Ma or ca. 124 Ma, depending on which age model for the Early Cretaceous is used.

The well-known dinosaur footprints of the Helvetiafjellet Formation at Festningen are constrained to the middle Barremian and, coupled with floral data, support a warm late Barremian prior to the Aptian carbon-isotope event. The appearance of glendonites at 655 m in the Carolinefjellet Formation is consistent with global cooling in the late Aptian–early Albian.

INTRODUCTION

The Early Cretaceous was a period that saw extreme perturbations in the carbon cycle and, it is thought, global climate. Paleo–high-latitude sediments deposited at this time are of key importance to understanding these global geochemical and climate changes because the poles of the Earth are extremely sensitive to climate change (e.g., Holland and Bitz, 2003). The Barremian to Albian sedimentary deposits of Spitsbergen, part of the Svalbard archipelago located in the Arctic Ocean, include some of the paleolatitudinally highest dinosaur footprints (Hurum et al., 2006), evidence for temperate forests (Harland et al., 2007), paradoxically together with interpreted dropstones (Dalland, 1977), and enigmatic "cold water" glendonites (Kemper, 1987; Price and Nunn, 2010). However, due to the limited occurrence of biostratigraphically useful fossils in the Early Cretaceous succession of Spitsbergen, correlations between outcrops across the island have been difficult, and precise, high-resolution dating of this succession has not been possible. This has meant that fully interpreting the conflicting paleoclimatic evidence from this Early Cretaceous succession has not been possible in a global context; therefore, this paleo–high-latitude site has been of limited use for understanding Early Cretaceous global climate change.

In order to improve age constraints on this succession, a different method of dating the succession must be attempted, such as carbon-isotope stratigraphy. Large (>2‰) perturbations in the isotopic composition of carbon in marine carbonate and marine and terrestrial organic matter are recorded in Early Cretaceous sediments from around the globe and reflect changes in global carbon cycling. These well-documented perturbations may be used to correlate marine and terrestrial sections from around the globe (e.g., Jenkyns, 1995; Menegatti et al., 1998; Ando et al., 2002; Herrle et al., 2015). A number of studies have specifically used discrete, identifiable fragments of fossil plant material for their carbon-isotope stratigraphy (e.g., Gröcke et al., 1999; Heimhofer et al., 2003; Robinson and Hesselbo, 2004; Weissert and Erba, 2004; Gröcke et al., 2005; Jerrett et al., 2015) with the caveat that "only large [>2‰] reproducible patterns can be indicative of long-term shifts in atmospheric carbon-isotopic compositions" (Robinson and Hesselbo, 2004, p. 134).

In this study, we present new sedimentological and organic $\delta^{13}C_{wood}$ data from the Lower Cretaceous succession of Spitsbergen from a well-studied site known as Festningen. These data build upon the earlier $\delta^{13}C$ data across the J/K boundary of Hammer et al. (2012) and Koevoets et al. (2016) and will be used to (1) improve age constraints on the succession; and (2) assess the timing and duration of purported warm and cold episodes in the Barremian–Aptian, based on the occurrence of warm and cold climate indicators (including flora and fauna) at the Festningen locality. These data will be integrated with those from previous studies of the Lower Cretaceous of Spitsbergen and will further be compared to the timing of such events in the Lower Cretaceous elsewhere.

GEOLOGICAL SETTING AND PREVIOUS WORK

The Svalbard archipelago is part of the greater Barents Sea region, located between 74 °N and 81 °N on the northwestern corner of the Barents Shelf. The principal island of Spitsbergen is situated at ~78 °N. During the Early Cretaceous, Spitsbergen was located at ~60 °N (Fig. 1; Torsvik et al., 2002;

Blakey, 2011) and was part of a shallow epicontinental sea that formed during the Mesozoic as Atlantic rifting propagated northwards. The Lower Cretaceous sedimentary succession on Spitsbergen consists of a regressive-transgressive megacycle and is divided into the broadly Berriasian–Barremian Rurikfjellet Formation, the Barremian Helvetiafjellet Formation, and the Aptian–Albian Carolinefjellet Formation. Pulses of volcanic activity, related to the emplacement of the High Arctic Large Igneous Province (HALIP) occurred during the Early Cretaceous on Svalbard (Harland, 1997; Maher, 2001; Nejbert et al., 2011; Senger et al., 2014; Polteau et al., 2016). The complete subaerial exposure of Svalbard during the Late Cretaceous has resulted in no Upper Cretaceous sediments being preserved on Svalbard (Harland, 1997).

The lowermost Rurikfjellet Formation is composed of two members: the Wimanfjellet Member containing the Myklegardfjellet Bed at its base and the overlying Kikutodden Member (Mørk et al., 1999; Midtkandal et al., 2008). This formation consists of mudstones and sandstones comprising several coarsening-upward successions deposited in marine shelf to prodeltaic environments (Dypvik et al., 1991). The abrupt contact of the Rurikfjellet Formation with the overlying Helvetiafjellet Formation is linked to thermal doming and a pulse of igneous activity associated with HALIP emplacement (Maher, 2001; Polteau et al., 2016). The Helvetiafjellet Formation was deposited within a broad fluviodeltaic system, with high sand influx, which gradually retreated landward due to rising sea level (Gjelberg and Steel, 1995; Midtkandal and Nystuen, 2009). The Helvetiafjellet Formation grades conformably into the overlying Carolinefjellet Formation, both formations recording the marine transgression that characterizes the upper Lower Cretaceous package on Spitsbergen (Nagy, 1970; Mørk and Worsley, 2006). This transgression is thought to result, in part, from eustatic sea-level rise (e.g., Haq et al., 1987; Harland, 1997). The Carolinefjellet Formation consists of five members, defined based on the predominance of sandstones or shales: the Dalkjegla Member, Innkjegla Member, Langstakken Member, Zillerberget Member, and Schönrockfjellet Member. Only the Dalkjegla and Innkjegla members are formally defined (Mørk et al., 1999). Commonly only the first two or three members (Dalkjegla, Innkjegla, and Langstakken members) are exposed in the northerly outcrops; the uppermost Schönrockfjellet Member only outcrops locally on south-easternmost Spitsbergen (Mørk et al., 1999).

Some authors (e.g., Gjelberg and Steel, 1995; Midtkandal et al., 2008) suggest that a major subaerial unconformity separates the Rurikfjellet Formation from the Helvetiafjellet Formation at some northwesterly locations, although the Rurikfjellet Formation grades conformably upwards into the Helvetiafjellet Formation in the southeast of Spitsbergen. Whether the uplift and erosion was great enough to cause a significant disconformity between the Rurikfjellet Formation and Helvetiafjellet Formation in the northwest is unclear (see Grøsfjeld, 1992).

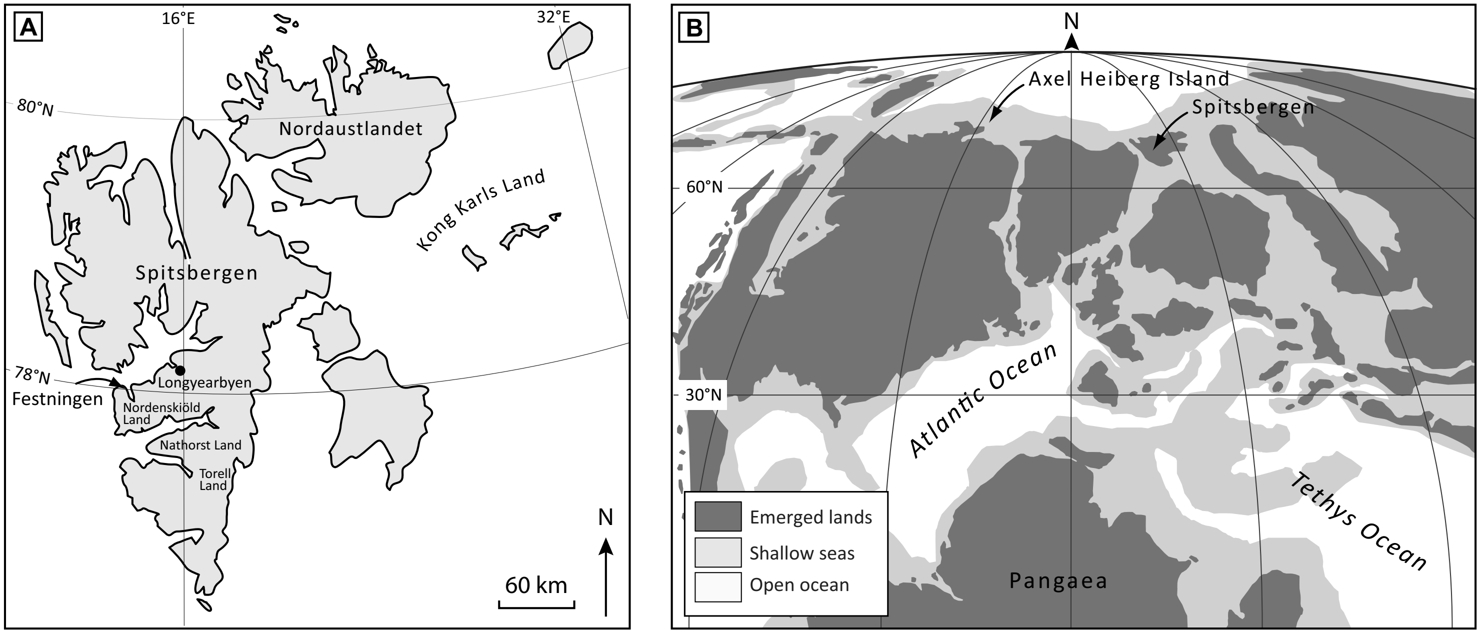

Figure 1. (A) Location of Festningen on Spitsbergen and other locations mentioned in the text. (B) Early Cretaceous paleogeographic reconstruction after Blakey (2011).

Age Model for the Festningen Section

The current age model for the formations making up the Festningen section is based on a combination of biostratigraphy and lithostratigraphy (Frebold, 1928; Hoel and Ørvin, 1937; Pčelina, 1970; Harland, 1972; Dypvik, 1984; Århus, 1992; Grøsfjeld, 1992; see Fig. 2 for an age model of the Spitsbergen Lower Cretaceous succession). Early studies suggest a Berriasian to Valanginian/Early Hauterivian age for the Rurikfjellet Formation at Festningen (as summarized by Frebold [1928] and Hoel and Ørvin [1937]). Later palynological studies from the Adventdalen region suggest that the Rurikfjellet Formation extends into the Barremian toward the top of the succession (Dypvik, 1984; Grøsfjeld, 1992). The amount of erosion of the Helvetiafjellet Formation into the Rurikfjellet Formation is variable across Spitsbergen; thus, the age of the upper Rurikfjellet Formation at Festningen can only be estimated based on lithostratigraphic correlations, which can certainly be improved upon. The Helvetiafjellet Formation has a paucity of biostratigraphically useful fossils in all studied outcrops and has therefore been dated only indirectly, using lithostratigraphic relations and floral assemblages (Pčelina, 1970; Vasilevskaya, 1980, 1986; Grøsfjeld, 1992; Mørk et al., 1999). A broadly Barremian age has been assigned to the Helvetiafjellet Formation, possibly extending into the early Aptian. Biostratigraphy (ammonites and dinocysts from the Adventdalen region and other southern Spitsbergen locations; Parker, 1967; Nagy, 1970; Pčelina, 1977; Århus, 1992) gives an Aptian–Albian age for the Carolinefjellet Formation. At Festningen, which is located in the northwest, it is likely that the top of the Carolinefjellet Formation is older because the north of Spitsbergen experienced more uplift and erosion than the south during the Late Cretaceous. Most workers place the Barremian–Aptian boundary at around the Helvetiafjellet-Carolinefjellet Formation boundary (see Mørk et al., 1999), in the absence of any direct constraints on age in the Helvetiafjellet Formation. A bentonite (felsic tuff) has been noted at several localities around Spitsbergen, including at Festningen, occurring just below the Helvetiafjellet-Carolinefjellet Formation boundary (e.g., Parker, 1967; Corfu et al., 2013; Midtkandal et al., 2016). Corfu et al. (2013) dated a felsic tuff in a core taken from DH3 borehole near Longyearbyen, yielding a U-Pb age of 123.3 ± 0.2 Ma.

METHODOLOGY

Field Sampling

Samples and sedimentological data were collected from Festningen, located on the southwestern side of Isfjorden, on the island of Spitsbergen, part of the Svalbard archipelago (78°09.98′N, 13°94.32′E) (Fig. 1), in August 2014. Festningen has historically been used by geologists to study the Early Cretaceous succession on Svalbard, as it is vertically bedded, well exposed, and accessible. The entire Lower Cretaceous sedimentary succession is exposed at Festningen (around 800 m thick), although macro-wood fossils are only found between the Kikutodden Member in the Rurikfjellet Formation and the lower half of the Innkjegla Member in the Carolinefjellet Formation. Thus for this study, the section between 396 m and 706 m was conventionally logged at a scale of 1 m = 2 cm, recording sedimentary structures, body, and trace fossils, and full range of grain sizes. Macro-plant material (ranging from coal to charcoal [Fig. 3]; see Fig. 4 for photographs of in situ plant material) was sampled wherever it occurred with high-resolution sampling possible from the top of the Rurikfjellet Formation, throughout the Helvetiafjellet Formation, and into the lower Carolinefjellet Formation. Bulk rock (mudstone-siltstones, organic rich) was also sampled every 0.5 m, where possible. In total, 139 macro-wood samples and 104 bulk-rock samples were collected. Bulk samples were taken by excavating the surface rock by up to 30 cm in order to mitigate the effects of surface weathering on subsequent analysis.

Microscope and Scanning Electron Microscope Wood Analysis

Different mechanisms may lead to the preservation of organic material in plants, and these may potentially affect the $\delta^{13}C$ recorded in their fossils. Ideally discrete, identifiable components should be chosen when analyzing terrestrial carbon isotopes (Heimhofer et al., 2003; Robinson and Hesselbo, 2004; Gröcke et al., 2005).

The preservational state (coal, wood, charcoal, and homogenized charcoal-coal) of the discrete fragments of woody material collected for this study was determined by examination of each sample under a light microscope (following the method of Jones and Chaloner [1991] and Gröcke et al. [2005]), in combination with high-resolution examination of a representative selection using a scanning electron microscope (SEM; Jeol JSM 6610LV SEM; see Fig. 3). This was done so that the carbon-isotope results could be assessed for any bias resulting from the different preservational states.

$\delta^{13}C$ and Total Organic Carbon Analysis

A total of 139 samples of coal, wood, charcoal, and homogenized charcoal-coal were analyzed for carbon-isotopic composition. These samples were ground to a fine powder using an agate mortar and pestle. Powdered samples were decarbonated by placing the sample in a 50 ml polypropylene centrifuge tube and treating with 10% HCl for 1 h until all the carbonate had reacted. Samples were then rinsed with deionized water, centrifuged, and rinsed again until neutrality was reached (following the method of Gröcke et al., 1999).

Carbon-isotope analysis was performed at Plymouth University using an Isoprime isotope ratio mass spectrometer connected to an Isoprime MicroCube elemental analyzer. Between 0.01 and 1.70 mg of sample were measured into tin capsules for analysis. Carbon-isotope ratios are expressed in the internationally accepted per mil (‰) standard notation relative to the Vienna PeeDee belemnite (VPDB) standard. Instrument calibration was achieved using two international standards: U.S. Geological Survey (USGS) 40 (l-glutamic acid, $\delta^{13}C$ = –26.389‰) and USGS 24 (graphite, $\delta^{13}C$ = –16.049‰). The mean $\delta^{13}C$ value and the standard deviation on replicate in run analyses of USGS 40 standard was 26.49 ± 0.08‰.

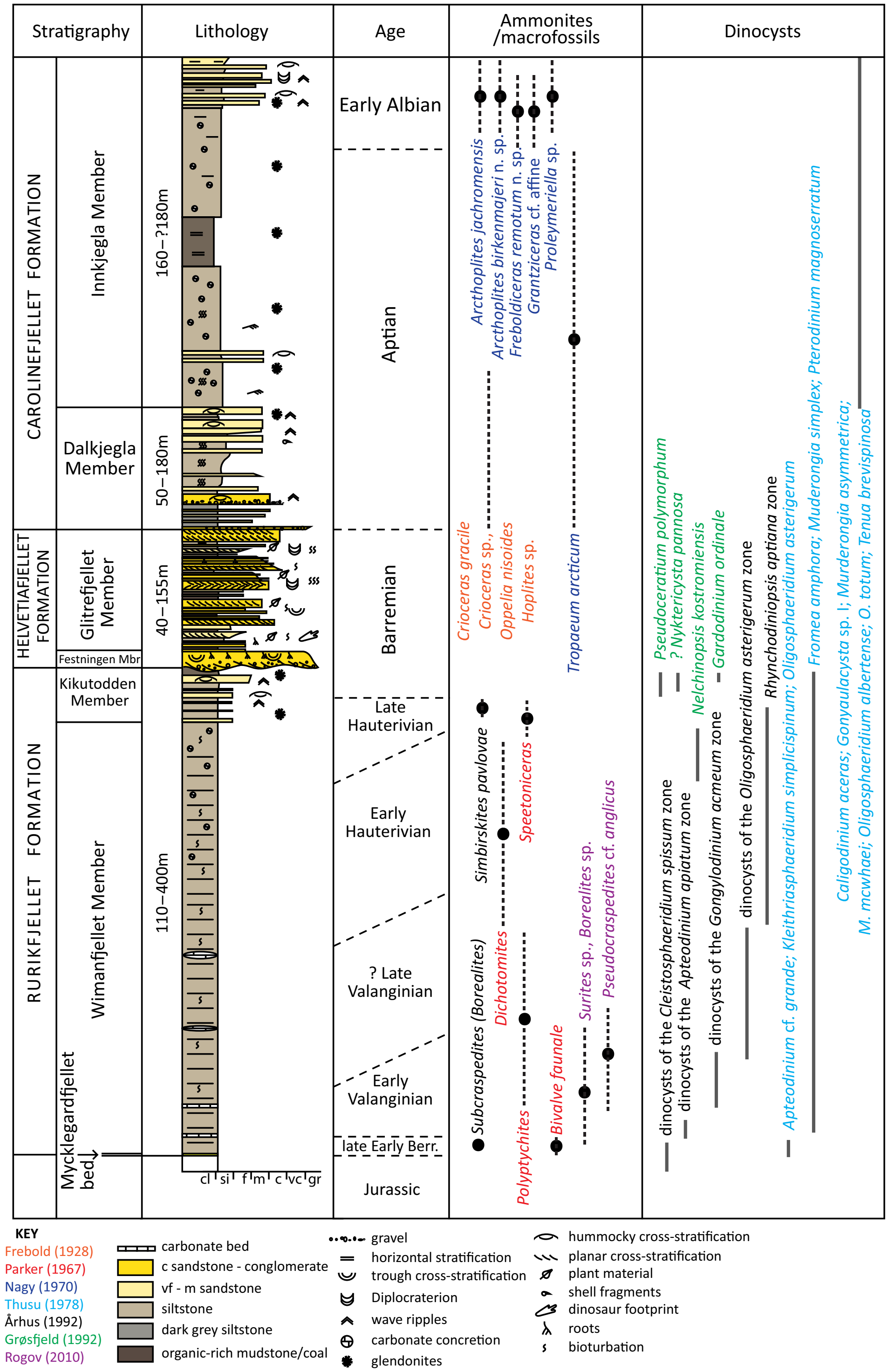

Figure 2. Age model for the Early Cretaceous succession on Spitsbergen showing previous ammonite and dinocyst data relative to a composite section representing the Lower Cretaceous succession of Spitsbergen (Frebold, 1928; Parker, 1967; Nagy, 1970; Thusu, 1978; Århus, 1992; Grøsfjeld, 1992; Rogov, 2010). Dashed lines indicate uncertainty range as to which stratigraphic level the samples were collected from in relation to the composite stratigraphic section. Composite stratigraphic section adapted from LoCrA Consortium (2013), with additional data from Maher et al. (2004) and field observations. Thickness ranges from Mørk et al. (1999) and LoCrA Consortium (2013).

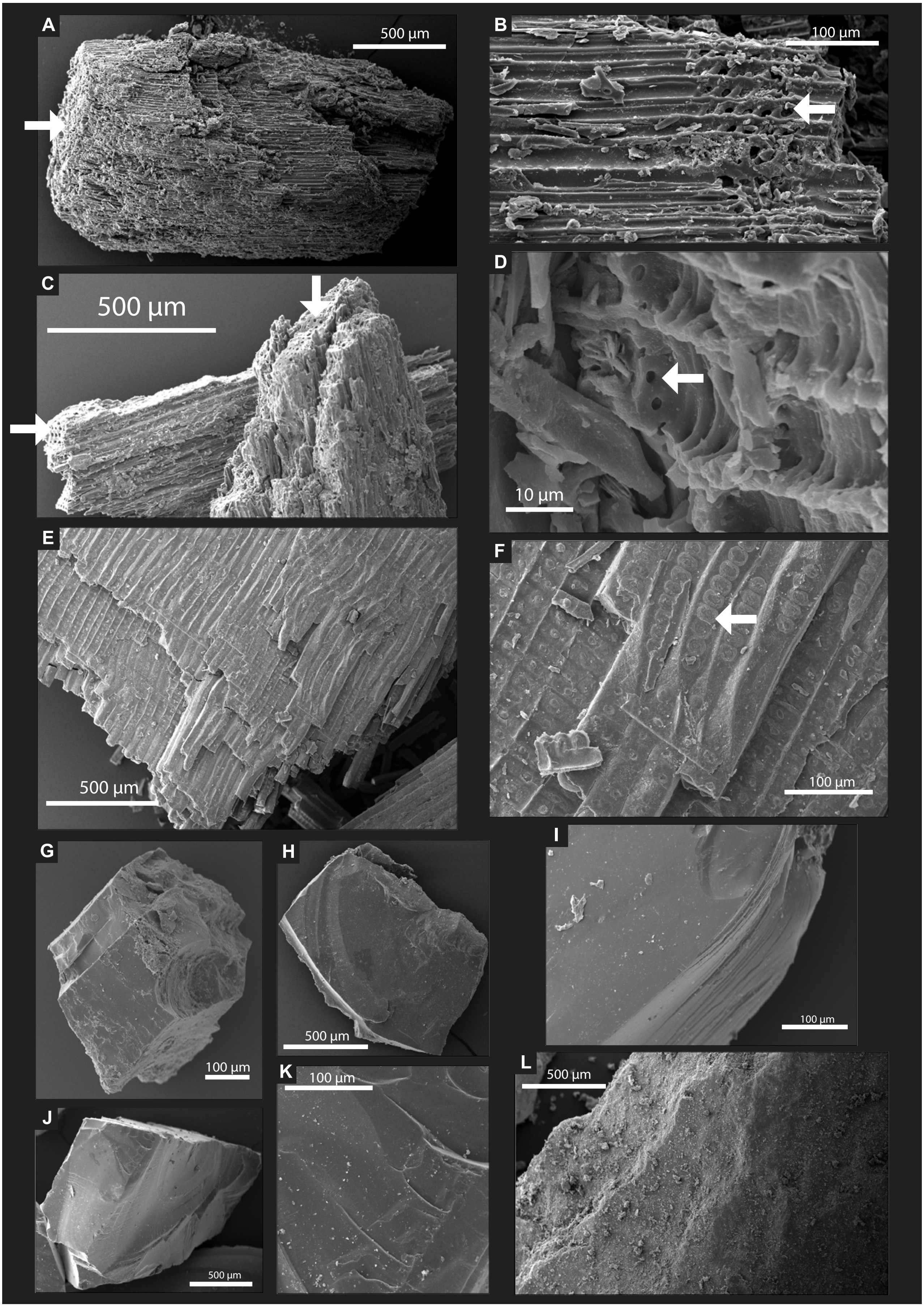

Figure 3. Scanning electron microscope photomicrographs illustrating the different preservation states found in the woody material from Festningen. (A–D) Charcoal. Blocky fragments; wood grain and cellular layout perceptible, although cell walls have become homogenized. Arrows point to porosity. (E and F) Non-coalified fossil wood. Cell structure visible (arrow points to cell). Wood has not been homogenized. (G–K) Coalified wood. Wood has been completely homogenized; no cell texture visible. Conchoidal fracturing of blocks. (L) Homogenized charcoal-coal. Total homogenization of cells. Note sediment on surface. Blocky fragments.

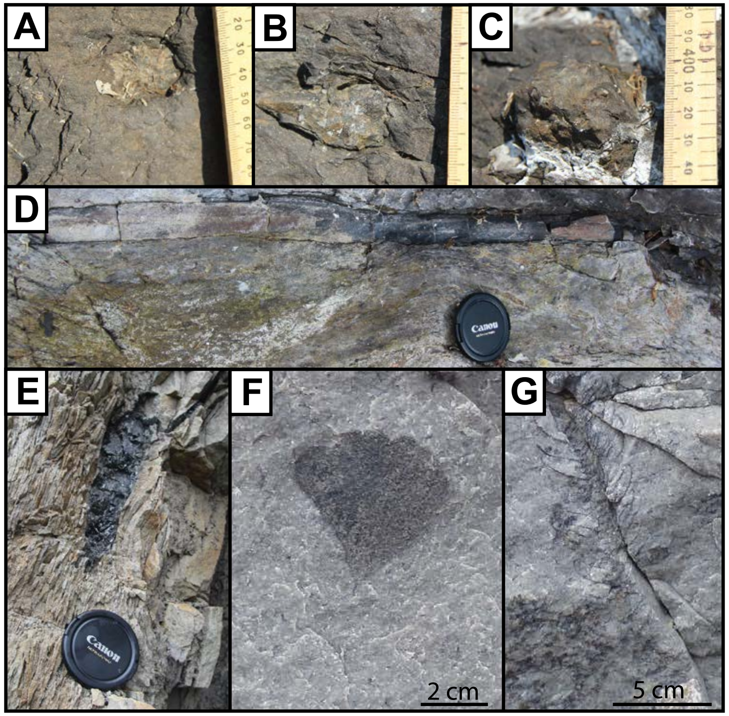

Figure 4. Field photographs of in situ glendonites (A–C); wood (D); coal (E); Ginkoales leaf impression (F); and Equisitales leaf impression (G) from Festningen.

Total organic carbon (TOC) analysis was performed on 104 ground bulk-rock samples using a Skalar Primacs solid and/or liquid carbon (SLC) analyzer (CS22), with a reproducibility of 0.3% based on repeat analyses of the same sample.

RESULTS

Description of Log

A summary log of the studied part of the Lower Cretaceous succession at Festningen is shown in Figure 5. The sediments of the upper Rurikfjellet Formation are composed of predominantly mudstones with sandstones comprising several coarsening-upward successions (consistent with previous observations suggestive of a regressive succession deposited above the storm wave base, e.g., Mørk et al., 1999). Lenticular concretions up to a meter in diameter together with cannon-ball–shaped nodules of siderite are particularly common. The thick sandstone package defining the Festningen Sandstone Member of the Helvetiafjellet Formation follows, resting with a marked erosional contact (at 453 m) on the underlying Rurikfjellet Formation. At Festningen, the lowermost 10-m-thick Festningen Sandstone Member consists of two erosively based, trough–cross-stratified, coarse sandstone units separated by a thin unit of finer-grained organic-rich shale with fine sandstone lenticles. Thin coal-bearing silt beds are found throughout the second sandstone package. The Festningen Sandstone Member is hard and tends to weather proud of the underlying units; it is thus useful for identifying the boundary between the Rurikfjellet and Helvetiafjellet formations.

Plant- and coal-bearing marine-to-paralic heterolithic sediments of the Glitrefjellet Member overlie the Festningen Sandstone Member. The sand content decreases upward through this member, and the Helvetiafjellet Formation grades conformably into the overlying Carolinefjellet Formation, with both formations recording the marine transgression that characterizes the upper Lower Cretaceous package on Spitsbergen (see Nagy, 1970; Mørk and Worsley, 2006). There is a marked color change toward the top of the Helvetiafjellet Formation succession (from pale gray at the base to grayish-green at the top) due to volcanic material (most likely sourced from the HALIP; Maher, 2001) mixing with the prevailing deltaic sediments (Worsley et al., 1986). Biostratigraphically useful ammonites were looked for but not found in the studied section of the succession at Festningen.

Macro-wood fossils were first identified at 402 m and were found scattered throughout the succession up until 691 m. Glendonite horizons were identified in the Kikutodden Member of the Rurikfjellet Formation at 408 m, 411.5 m, 441 m, and in the Innkjegla Member of the Carolinefjellet Formation, at 655 m and 699.5 m. Dinosaur footprints, as well as impressions of leaves of the Equisitales and Ginkoales orders, were observed between 469 and 471 m. See Figure 4 for field photographs of glendonites and fossil plant material. The preservation state of the macro-wood samples ranged from charcoal to coal. In total, 14% of the samples were classified as coal, 2% as wood, 77% as homogenized charcoal-coal, and 7% as charcoal. The group categorized as homogenized charcoal-coal consists of totally homogeneous organic material, which cannot be identified further (see Fig. 3).

Description of $\delta^{13}C$ and Total Organic Carbon Data

The $\delta^{13}C_{wood}$ mean and standard deviation for the entire population are –24.3‰ and 0.90‰, respectively. The mean value for the 21 coalified fragments is –24.4‰. The mean for homogenized charcoal-coal is –24.3‰; that for charcoal is –24.0‰; and that for wood is –23.9‰. The carbon-isotope results for the woody material are plotted in Figure 5.

The most negative $\delta^{13}C_{wood}$ value is –27.2‰ from sample 535 m; the most positive $\delta^{13}C_{wood}$ value is –22.0‰ from sample 563.5 m, and the overall variation in $\delta^{13}C_{wood}$ is 5.2‰. There is a positive shift at 504.4 m from –25.0‰ up to –22.4‰, with a return to more negative values (–25.15‰) at 510.6 m. A pro-

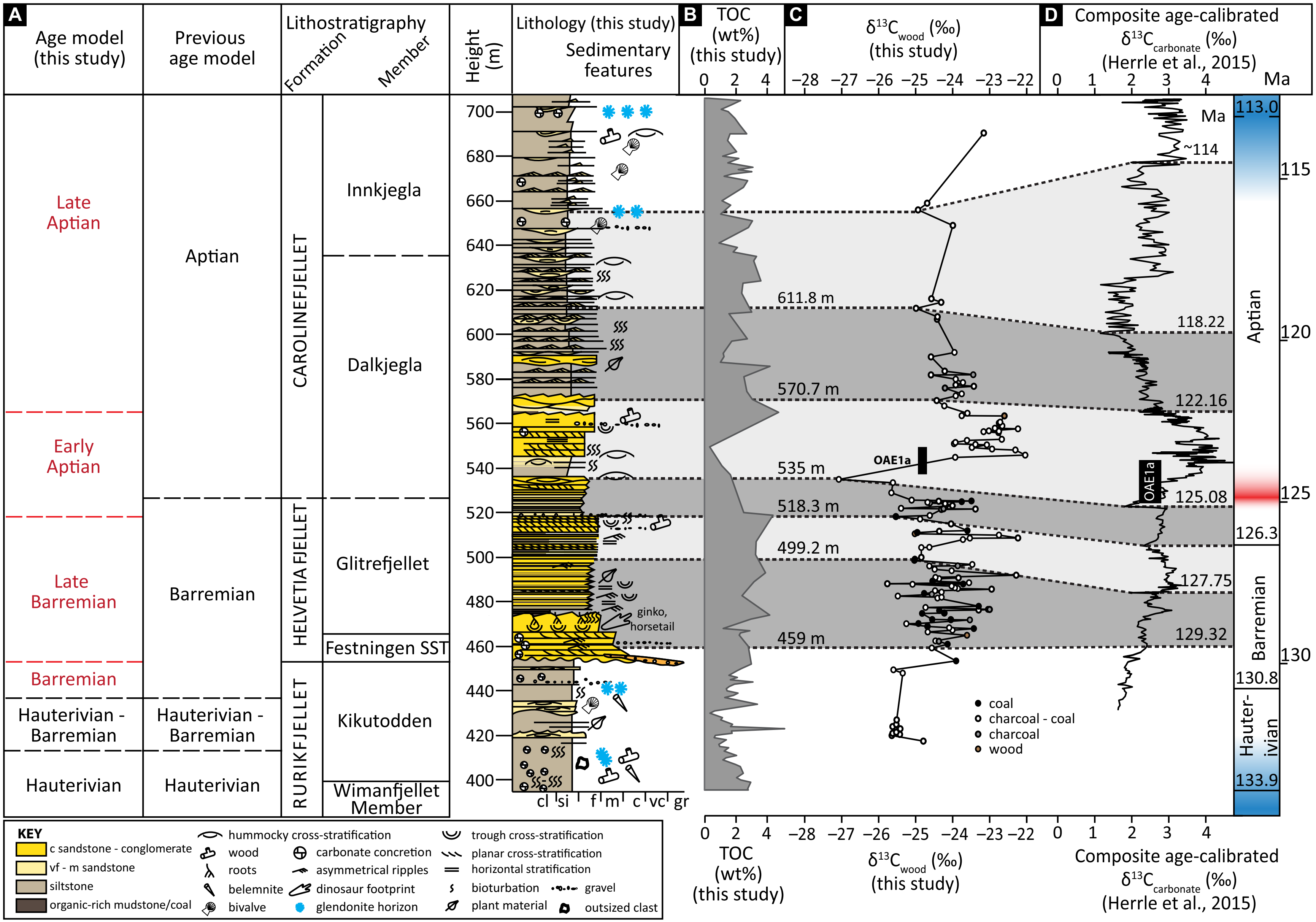

Figure 5. Sedimentary succession and δ13Cwood record at Festningen, Spitsbergen. (A) Stratigraphy, lithostratigraphy, and lithology from field observations. (B) Total organic carbon (TOC) for shales interbedded with sandstones throughout the succession. (C) δ13C results for discrete wood fossils (ranging from charcoal to coal). (D) Composite age-calibrated δ13C curve (from Herrle et al., 2015; using the timescale of Ogg and Hinnov, 2012). Dark- and light-gray areas indicate correlative intervals. Global cool episodes indicated by blue shaded area on timescale; warming episodes indicated in red shaded area on timescale. Position of Aptian Ocean Anoxic Event (OAE1a) relative to the carbon-isotope records also displayed. Festningen SST—Festningen Sandstone.

metres	δ13Cwood	type		type
417.50	-24.92	2		1 coal
420.00	-25.77	2		2 homogenised charcoal-coal
420.50	-25.55	2		3 charcoal
421.00	-25.75	2		4 wood
421.50	-25.65	2		
423.00	-25.54	2		
424.00	-25.75	2		
425.00	-25.65	2		
427.00	-25.64	2		
448.00	-25.48	2		
449.50	-25.73	2		
453.50	-24.03	1		
459.30	-24.69	2		
461.00	-24.26	1		
462.30	-24.49	3		
462.50	-24.59	2		
465.00	-23.73	4		
466.50	-24.80	2		
467.30	-24.13	2		

[1]Supplemental File 1. $\delta^{13}C_{wood}$ data. Please visit http://dx.doi.org/10.1130/GES01344.S1 or the full-text article on www.gsapubs.org to view Supplemental File 1.

FST height (m)	wt % TOC
396.0	2.75
399.0	2.95
400.0	2.31
401.0	2.04
403.0	2.47
406.0	2.28
407.0	1.97
409.0	1.16
410.0	1.37
411.0	2.04
413.0	1.24
414.5	1.53
415.0	0.96
416.0	0.90
418.0	1.38
419.0	0.60
420.0	0.67
421.0	1.75
422.0	1.71

[2]Supplemental File 2. TOC data. Please visit http://dx.doi.org/10.1130/GES01344.S2 or the full-text article on www.gsapubs.org to view Supplemental File 2.

nounced negative shift of >3‰ in $\delta^{13}C_{wood}$ at 535 m to –27.2‰ is observed, followed by a positive shift of >5‰ to –22.15‰ at 545.9 m. There is a second positive peak at 557.70‰ to –22.35‰. This is followed by a gradual decrease in $\delta^{13}C_{wood}$ values to 611.8 m. $\delta^{13}C_{wood}$ data are listed in Supplemental File 1[1].

The TOC mean and standard deviation for the population of bulk rock samples are 2.10% and 0.97%, respectively. The range of values for the population is 4.89%. These TOC results are also plotted in Figure 5. TOC data are listed in Supplemental File 2[2].

DISCUSSION

The $\delta^{13}C_{wood}$ curve from the Festningen section records >3‰ carbon-isotope excursions also seen in the calibrated composite $\delta^{13}C_{carbonate}$ curve (reproduced from Herrle et al., 2015, using published $\delta^{13}C$ carbonate records from Tethyan sections of Erba et al., 1999; Herrle et al., 2004; Gale et al., 2011, and calibrated using Ogg and Hinnov, 2012). The means of the different categories of wood material are within 0.5‰ of each other. This leads us to conclude that the different preservational states of the samples have not significantly biased the $\delta^{13}C_{wood}$ data.

Furthermore, the fact that long-term trends seen in other carbon-isotope records (e.g., Jenkyns, 1995; Menegatti et al., 1998; Heimhofer et al., 2003; Herrle et al., 2015) are reproduced in this data set suggests that the changes in $\delta^{13}C_{wood}$ are controlled by global carbon cycle perturbations and not by local paleoenvironmental or paleoecological effects, and these trends support the utility of wood $\delta^{13}C$ series as a viable technique for dating deltaic or even terrestrial sections that may be biostratigraphically poor.

Carbon isotopic ratios ($\delta^{13}C$) from terrestrial (plant) organic material have been previously used to study global carbon-isotope excursions (CIEs) in the Cretaceous (e.g., Ando et al., 2002; Heimhofer et al., 2003; Robinson and Hesselbo, 2004; Gröcke et al., 2005; Herrle et al., 2015; Jerrett et al., 2015). These studies (which use either discrete plant fragments or disseminated T_{org} from well-dated sections) all record >3‰ excursions also found in the carbonate record showing that the preservational style of woody material does not influence the recorded $\delta^{13}C_{wood}$ when identifying large reproducible patterns in atmospheric carbon isotopic compositions, including Hesselbo et al. (2000), Hesselbo et al. (2003), Robinson and Hesselbo (2004), Gröcke et al. (2005), and Jerrett et al. (2015).

Chemostratigraphy

Using our chemostratigraphic correlation and $\delta^{13}C_{wood}$ curve with the composite global carbonate curve (Herrle et al., 2015), the early Aptian CIE can be identified across the Helvetiafjellet-Carolinefjellet Formation boundary. The early Aptian CIE has been associated with the Aptian Ocean Anoxic Event (OAE1a; e.g., Menegatti et al., 1998; Price, 2003; van Breugel et al., 2007), a time of increased burial of organic carbon and extreme global warmth (Schouten et al., 2003; Weissert and Erba, 2004; Dumitrescu et al., 2006). Global marine carbonate, marine, and terrestrial organic carbon records from the Aptian all record similar $\delta^{13}C$ perturbations to that seen in this study, whereby a sharp negative excursion at the onset of OAE1a (Selli level; early Aptian, of –4.0‰ to –7.5‰ depending on which carbon record is being examined; Heimhofer et al., 2003) is followed by two positive excursions in the mid-Aptian (e.g., Menegatti et al., 1998; Jones and Jenkyns, 2001; Ando et al., 2002; Heimhofer et al., 2003; Price, 2003; Weissert and Erba, 2004; Keller et al., 2011). The prominent short-term negative carbon excursion (Fig. 5; Menegatti et al., 1998) has been attributed to either the addition of ^{13}C-depleted carbon to the ocean-atmosphere reservoir from marine volcanism (e.g., Bralower et al., 1994, 1999) or dissociation of methane gas hydrates (e.g., Jahren et al., 2001; van Breugel et al., 2007), both mechanisms which could result in extreme global warming.

The $\delta^{13}C$ curve from the Festningen wood samples in this study shows the sharp negative excursion (>3‰) followed by a double-positive (~5‰) spike. The negative spike is defined by four $\delta^{13}C$ data points, with the most negative point at 535 m. A significant double-positive excursion follows, defined by 23 data points. The shape of the curve across this interval closely matches that of the composite carbonate curve, as well as other published carbonate and organic carbon curves, e.g., Jenkyns (1995). This leads us to conclude that this section of the Festningen succession does indeed record the OAE1a CIE. The sediments here represent the lower part of the Dalkjegla Member (of the Carolinefjellet Formation) and consist of heterolithic sands and muds, coarsening upwards (and increasing in sand content) to 545 m (Fig. 5). This package has TOC values of up to 5% and is overlain by an organic-poor (<1% TOC) thick sand package. Thus within this part of Svalbard, the early Aptian CIE occurs in a transgressive unit comparable with findings from the Canadian Arctic (Herrle et al., 2015) and the Tethyan Vocontian Basin in France (Herrle et al., 2004).

Our $\delta^{13}C$ chemostratigraphic approach allows us to evaluate the age of the Rurikfjellet-Helvetiafjellet Formation boundary. An abrupt regressive contact at the base Helvetiafjellet Formation is observed at Festningen. This erosive contact of the Helvetiafjellet Formation into the Upper Rurikfjellet Formation has been noted at other localities across Spitsbergen (see Fig. 1), from Nordenskiöld Land to the east of Festningen (e.g., Midtkandal et al., 2007), Lardyfjellet (even farther east, e.g., Dypvik et al., 2002), to Nathorst Land, and down to Torell Land to the south (see Midtkandal et al., 2008). The uplift that resulted in the subaerial exposure of the Rurikfjellet Formation may be the result of a pulse of volcanic activity, related to the emplacement of the HALIP (Harland, 1997; Maher, 2001; Nejbert et al., 2011, Senger et al., 2014; Polteau et al., 2016).

Using our $\delta^{13}C$ curve correlation to the composite age-calibrated $\delta^{13}C$ carbonate curve (Ogg and Hinnov, 2012), we estimate that the Rurikfjellet-Helvetiafjellet Formation boundary occurs at ca. 129 Ma. Such an age estimate is not inconsistent with published geochronologies, which indicate magmatism beginning at 135 Ma (Maher, 2001) or 125.5 ± 3.6 Ma (Nejbert et al., 2011).

However, there is considerable debate as to the absolute ages for the Early Cretaceous, particularly in the Barremian and Aptian. Other age models (e.g., He et al., 2008; Malinverno et al., 2010) suggest that M0r (which defines the Barremian–Aptian boundary) occurs at 121.2 ± 0.5 Ma, some 5 m.y. younger than the ages published by Ogg and Hinnov (2012).

Other geochronological data (e.g., U-Pb detrital-zircon ages from the Southern Patagonian Andes; Ghiglione et al., 2015) are in agreement with the base age of the Aptian calculated by He et al. (2008). If this is so, the Rurikfjellet/Helvetiafjellet Formation may in fact be closer to 124 Ma. This age still remains consistent with the timing of magmatism as published by Nejbert et al. (2011) and Senger et al. (2014), and more closely agrees with timing of magmatism on Svalbard as published by Polteau et al. (2016).

Our $\delta^{13}C_{wood}$ data set may also be used to constrain the location of the Barremian–Aptian boundary in the Festningen section, as well as age of the boundary. Despite uncertainties regarding the absolute age of this boundary (as noted above), it is widely accepted to occur at M0r, which itself is below the large negative CIE marking the Selli level. The Barremian–Aptian boundary is marked by a lesser negative excursion (e.g., Menegatti et al., 1998; Ogg and Hinnov, 2012; Herrle et al., 2015). Considering the position of the large negative excursion, we correlate the small negative excursion at 518.3 m (17 m below the "Selli level" data point), to the Barremian–Aptian boundary on the composite $\delta^{13}C$ carbonate curve. The U-Pb zircon age dated tuff of the Helvetiafjellet Formation at Adventdalen (Corfu et al., 2013; Polteau et al., 2016) should be located in the Festningen succession just below the large negative excursion (close to the top of the Barremian). Using the Ogg and Hinnov (2012) age model, the tuff dated as 123.3 ± 0.2 Ma (Corfu et al., 2013) would occur at the end of the early Aptian, whereas if He et al.'s (2008) model is used (with a Barremian–Aptian boundary age of 121.2 Ma), then this tuff would be within the late Barremian. This highlights the need for a better age constraint for the Barremian–Aptian boundary. Indeed, recently, Midtkandal et al. (2016) suggest that it be revised to a younger age 121–122 Ma, in agreement with the age model of He et al. (2008).

Record of Climate Change in Svalbard

Our detailed stratigraphic framework also enables us to examine the timing and significance of Arctic faunal and floral data. Previously, dinosaur footprints reported at Festningen and Kvalvågen were thought to have been broadly Barremian in age (e.g., Harland, 1997; Hurum et al., 2006). Our new data constrain the dinosaur footprints of Festningen to the middle Barremian. Indeed, temperatures throughout the late Barremian are thought to have been fairly warm (e.g., Mutterlose et al., 2010) and if extrapolated poleward would imply conditions suitable to sustain large ectotherms. The flora (e.g., Gingkoales and Equisitales) preserved in the Helvetiafjellet Formation at Festningen likewise suggest a warm environment—but humid rather than arid (e.g., Hurum et al., 2006).

Following this period of warmth, the earliest Aptian experienced a brief, intense warming event, associated with OAE1a (thought to be due to the release of methane or large amounts of volcanic CO_2 into the atmosphere; e.g., Bralower et al., 1994, 1999; Jahren et al., 2001; Weissert and Erba, 2004; van Breugel et al., 2007). Cooling succeeded this event, as enhanced carbon burial sequestered the excess CO_2, reflected by a positive shift of oxygen-isotope records from Tethyan regions during the late early Aptian (Hochuli et al., 1999; Keller et al., 2011; Bodin et al., 2015). However, there is no direct evidence of extreme warming or cooling at this time in the sediments preserved at Festningen.

A second cooling event is thought to have occurred in the late Aptian, as a number of different proxies have indicated, e.g., $\delta^{18}O$, TEX_{86}, calcareous nannofossils (e.g., Herrle and Mutterlose, 2003; Mutterlose et al., 2009; McAnena et al., 2013; Bodin et al., 2015), and this is not inconsistent with our findings at Festningen. The first appearance of glendonites, which may be associated with cold water conditions (as well as specific chemical conditions; see Bischoff et al., 1993; De Lurio and Frakes, 1999; Rickaby et al., 2006; Selleck et al., 2007; and Zhou et al., 2015, for more detail), in the Carolinefjellet Formation at Festningen occurs at 655 m. Although macro-wood fossils above 580 m are sparse, so that using carbon-isotope stratigraphy to date their appearance more accurately is not possible, previous studies suggest that the Inkkjegla Member (in which the Carolinefjellet Formation glendonites occur) is late Aptian–early Albian in age (e.g., Nagy, 1970; Thusu, 1978; see Fig. 2). Herrle et al. (2015) also report the occurrence of late Aptian–earliest Albian glendonites from Axel Heiberg Island, Canada, coeval with a subtropical Atlantic drop in sea surface temperatures of ~4 °C (McAnena et al., 2013). This fits with floral data from Harland et al. (2007), who suggest that Svalbard was cool-temperate during the Aptian–Albian with temperatures varying between 1 and 10 °C, based on the presence of *Taxaceoxylon* in the Innkjegla Member in the Longyearbyen region of Spitsbergen. Thus it appears that the late Aptian–early Albian cooling event may be recorded in the sediments of Svalbard.

CONCLUSIONS

• Our terrestrial organic carbon-isotope record (macro-wood fossils) from Festningen shows a 3‰ negative excursion followed by a ~5‰ double-positive excursion, correlatable with the global early Aptian carbon-isotope event. This allows us to constrain ages for certain parts of the succession and assess the timing and significance of climatic indicators, such as glendonites, flora, and fauna.

• We provide an age estimate for the Rurikfjellet-Helvetiafjellet Formation boundary of ca. 129 Ma (sensu Ogg and Hinnov, 2012) or ca. 124 Ma (sensu He et al., 2008), both of which are consistent with published geochronologies that indicate a pulse of HALIP magmatism beginning at 135 Ma–125.5 ± 3.6 Ma (Maher, 2001; Nejbert et al., 2011; Senger et al., 2014; Polteau et al., 2016).

• Using our chemostratigraphic correlation, we identify the Barremian–Aptian boundary at 518.3 m, which lies toward the top of the Glitrefjellet Member of the Helvetiafjellet Formation. This, when looked at with bentonite age estimates from previous studies, highlights the need to revise the age assigned to the Barremian–Aptian boundary (as published in Ogg and Hinnov, 2012).

• Dinosaur footprints, observed at 465 m, along with impressions of leaves from the orders Ginkoales and Equisitales (consistent with a warm, humid climate), can now be constrained to the middle Barremian. The extreme global warming event followed by cooling that is thought to have accompanied the early Aptian carbon-isotope event is not strongly evident at Festningen. However, the appearance of glendonites in the Carolinefjellet Formation is consistent with global cooling in the late Aptian–early Albian. The presence of these glendonites support floral data from the Innkjegla Member studied by Harland et al. (2007), which suggests cool temperatures in the late Aptian–early Albian.

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